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CLIMATE-CONTROLLED GLACIAL EROSION IN THE UNCONSOLIDATED SEDIMENTS OF NORTHWESTERN EUROPE, BASED ON A GENETIC MODEL FOR TUNNEL VALLEY FORMATION

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ABSTRACT

The development of large erosive subglacial forms in unconsolidated sediments is generally attributed to the eroding power of subglacial meltwater flowing under high pressure conditions. Most explanations, however, differ in the source of meltwater and the speed at which it erodes the subglacial bed. Based on the geometry of deep tunnel valleys and glacial basins in northwestern Europe, a reconstruction of subglacial hydrological conditions during the development of subglacial depressions is made. It is demonstrated that the flow of subglacial meltwater in subglacial channels under high glaciostatic pressures is only capable of eroding large volumes of sediment as long as there is imminent glaciohydrological instability. For the thick aquifers in northwestern Europe, this instability is achieved when large quantities of supraglacial meltwater are available. Furthermore, a theoretical definition is given for maximum depression depth to be reached by subglacial erosion. It is shown that this maximum depth is strongly related to average air temperatures during deglaciation and that glacier bed lowering is to be expected during any final phase of glaciations. The theoretical framework presented enables a tentative comparison between large-scale glacial morphology of different glaciations in northwestern Europe.

KEY WORDS subglacial hydrology; deglaciation; Pleistocene ice sheets; Elsterian; Saalian; tunnel valley

INTRODUCTION

Reconstruction of the dynamics of the Scandinavian ice sheets during the Quaternary glacials in northwestern Europe is usually derived from the stratigraphical record which is preserved in the North Sea tectonic basin. Current knowledge of former maximum ice sheet extensions and rates of their retreat or advance is mainly based on the spatial distribution, patterns and interpretation of stratigraphical sequences of glaciogenetic sediments. Morphological effects of the advance of the Scandinavian ice sheets, however, are not restricted to depositional processes. There is ample evidence for glacial erosion which resulted in the widespread formation of tunnel valleys and glacial basins (Ehlers *et al.*, 1984). Correct genetic interpretation of these erosional forms may provide important additional information to current knowledge on the behaviour of large ice sheets overlying unconsolidated sediments.

Although glacial erosion took place during all three major glaciations in northwestern Europe, its effect on the landscape has differed from one glaciation to the other. It is even possible to characterize each of the last three glaciations by its resulting glacial morphology, as has been done by Ehlers (1983). Glacial erosion during the Elsterian age has resulted in the formation of very large tunnel valley systems, which cover much of northwestern Europe. This is in striking contrast to the glacial basin–push moraine landscape known from

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the Saalian. During the Weichselian glaciation, both tunnel valley and glacial basin formation took place, but to a significantly lesser extent and intensity than during the two previous glaciations.

Probably as a result of the contrast in glaciomorphological landscapes, most research aiming at a genetic interpretation of glacial erosive forms has focused on either tunnel valleys or glacial basins. This has led to a variety of explanations and interpretations of the behaviour of subglacial processes acting beneath large ice sheets overlying unconsolidated sediments. Generalization of these theories is often difficult as they are usually based on regional data. A major problem with respect to theory development is that it is seriously hampered by the lack of modern examples of large ice sheets overlying thick unconsolidated sediments.

There are, however, important indications that a genetic comparison between tunnel valleys and glacial basins could be promising. Both tunnel valleys and glacial basins are caused by glacial erosion beneath ice sheets which are comparable in size and are formed in approximately the same region. Many morphogenetic theories for both features are based on the same hydromechanical concepts. It is therefore surprising that little effort has been made so far to study the similarities between processes and morphometrical characteristics of the two erosional features.

It is the aim of this article to make a theoretical analysis of glacial erosion beneath a large ice sheet overlying the thick beds of unconsolidated sediments in northwestern Europe, and to derive most important factors controlling erosion. A model-supported interpretation of the erosional potentials of past glacial systems is made. Both field data and other models, described in the literature, are combined or adapted and applied. Besides a geometrical and sedimentological analysis of described Elsterian tunnel valleys and Saalian glacial basins in the literature, the application of theoretical glaciological concepts of earlier theories is evaluated and tested against available field evidence. Finally, an attempt is made to generalize the conclusions concerning physical conditions beneath ice sheets during the two major Quaternary glaciations in northwestern Europe.

GLACIAL MORPHOLOGY OF ELSTERIAN AND SAALIAN AGE IN NORTHWESTERN EUROPE

Because of our interest in possible similarities between the development of tunnel valleys and glacial basins, an analysis of the geometrical characteristics of these erosive forms may give a first insight into the conditions leading to glacial erosion. We therefore summarize the general characteristics of the Elsterian tunnel valley system and compare it with the Saalian glacial basin–push moraine landscape. This restriction is made because these landscapes can be regarded as extreme and 'clean' examples of the impact of one particular process, which have not been greatly disturbed by other glaciomorphological phenomena.

Elsterian tunnel valleys

Contour maps of Elsterian tunnel valley depths (Küster and Meyer, 1979; Hinsch, 1979; Bosch, 1991; de Groot, 1987; ter Wee, 1976, 1979; Cameron *et al.*, 1984; Jeffery *et al.*, 1989) reveal that dimensions of individual tunnel valleys show great variation. Elsterian tunnel valleys are about 100 to 200 m deep, with maximum depths of more than 400 m. Most tunnel valleys are 2 to 4 km wide and 30 to 100 km long (Figure 1).

The spatial pattern and geometry of the Elsterian tunnel valleys are characterized by two main aspects which distinguish them from ordinary fluvial valleys. Tunnel valley maps of northern Germany, The Netherlands and the North Sea indicate that tunnel valleys are not isolated forms but are often interconnected, displaying an anastomosing valley pattern. Their orientation is parallel to what is assumed to be the general ice flow direction of the Elsterian ice sheet (Ehlers *et al.*, 1984). The overall network pattern suggests that each tunnel valley is a unit of one uniform hydrological system, excluding any local origin of the valleys. The second aspect concerns the longitudinal profiles of individual tunnel valleys which are characterized by an irregular base with many thresholds (Ehlers and Linke, 1989; Ehlers, 1990; Piotrowski, 1994). Thus, tunnel valleys can be regarded as being composed of separate elongated depressions aligned end-to-end. Threshold altitudes can be up to 100 m. The observed discontinuity of the valley bottom gradient may be explained by both variable erodibility of the sediments or discontinuity of erosion activity beneath the ice sheet.

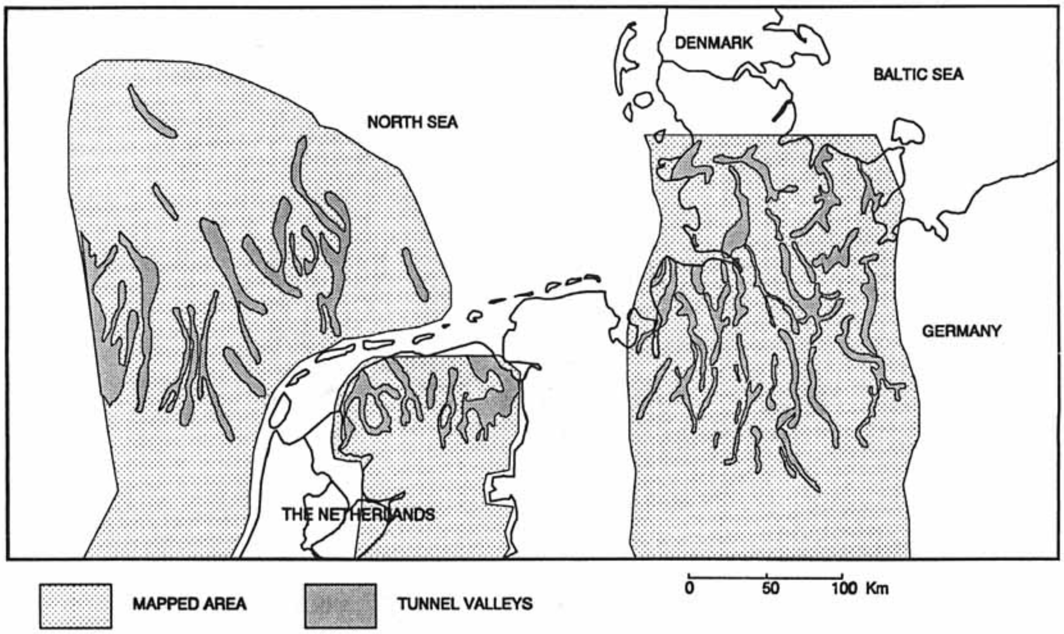


Figure 1. Compiled map of the occurrence and spatial geometry of Elsterian tunnel valleys in northwestern Europe, after Cameron *et al.* (1984), Stoker *et al.* (1985), Jeffery *et al.* (1989), ter Wee (1976, 1979), de Groot (1987), Bosch (1991), Hinsch (1979), Küster and Meyer (1979) and Ehlers (1990)

Infill sedimentology of the Elsterian tunnel valleys usually consists of an alternating sequence of meltwater deposits and glaciolacustrine sediments (Ehlers, 1990; ter Wee, 1976). Till has been found in some tunnel valleys (Linke and Ehlers, 1987; Piotrowski, 1994). Whereas the occurrence of till clearly points to glacial activity within the tunnel valleys, the implication of the occurrence of glaciolacustrine and meltwater deposits is most likely not related to the subglacial erosion process, but rather represents geomorphological processes within the vicinity of the tunnel valleys after ice sheet retreat.

Two types of genetic models have been proposed for the formation of deep tunnel valleys. Ehlers and Linke (1989), Ehlers (1990), Shoemaker (1992) and Wingfield (1990) proposed models for deep Elsterian tunnel valley development based on assumed catastrophic release of great volumes of sub- or supraglacially stored meltwater. A similar catastrophic model has been developed for the origin of (much smaller) tunnel valleys beneath the Superior Lobe during the Laurentidian ice age in Minnesota, USA (Wright, 1973). These catastrophic models are opposed by models where erosion is progressive and described as a function of steady-state meltwater flow in subglacial channels (Boulton and Hindmarsh, 1987; Mooers, 1989; Jeffery, 1991). In the steady-state drainage model, tunnel valley development is the result of prolonged downgrading of the bed of relatively small subglacial meltwater channels by meltwater erosion. This subglacial stream flows at the depression bottom while the valleys are filled with ice.

The advantage of the progressive models is that they adapt known glaciological processes and concepts to explain tunnel valley formation without the introduction of exceptional or even unrealistic circumstances. Unfortunately, the existing models are still unable to reconstruct the glaciological conditions which caused the extreme dimensions of Elsterian tunnel valleys of northwestern Europe in particular.

Saalian glacial basins

Glacial basins of the Saalian glaciation do not have a typical geometry, but are rarely deeper than 150 m below the former surface in which they are eroded. Their widths and lengths can extend up to several tens of kilometres (de Gans *et al.*, 1986; ter Wee, 1983; Jelgersma and Breeuwer, 1975; van der Wateren, 1985). Analysis of the spatial patterns of glacial basins in northwestern Europe clearly points to a relatively even distribution along a line over northwestern Europe in an east–west direction (Figure 2).

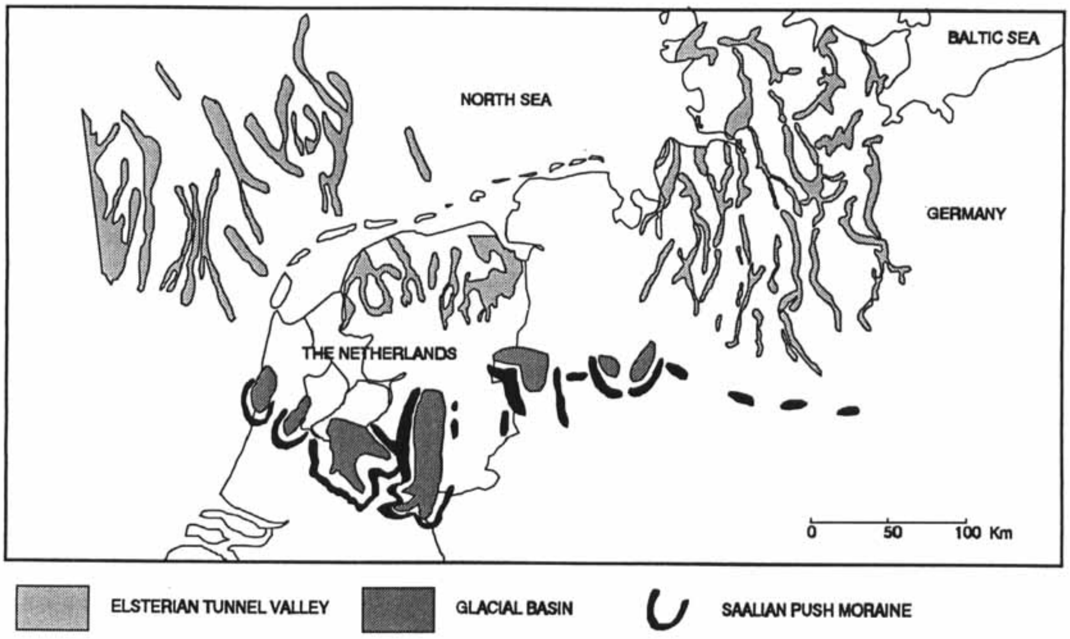


Figure 2. Schematic map showing the relation between the spatial distribution of Saalian glacial basins (with accompanying push moraines) and Elsterian tunnel valleys

Although the origin of glacial basins is relatively well understood, and explained as the removal of subglacial sediments caused by drag of the glacier bed due to differential ice load (Aber *et al.*, 1989), there is strong evidence that parts of the eroded basin sediments were transported by subglacial meltwater streams (van der Wateren, 1992). The effect of sediment transport by subglacial meltwater streams can be observed within the vicinity of the push moraine systems in northwestern Europe, which are characterized by the occurrence of large amounts of meltwater deposits in large proximal alluvial or outwash fan systems. Further evidence of the presence of subglacial meltwater is the occurrence of tunnel valleys at the margins of some glacial basins (de Gans *et al.*, 1986). Little attention has so far been given to the relative importance of the process of subglacial meltwater erosion preceding or occurring syngenetically with glacial basin formation.

MODELLING SUBGLACIAL HYDROLOGY

Based on the interpretation of field characteristics of geometry and sedimentology, it is clear that a dominant factor for both tunnel valley formation and glacial basin genesis is the role of subglacial meltwater. However, subglacial hydrology of Pleistocene glacier beds overlying unlithified sediments is complex and difficult to assess due to the lack of current examples and the inaccessibility of glacier beds. Available knowledge on subglacial hydrology derived from modern valley glaciers, of which many have hardrock beds, is hard to extrapolate because of the scale problem and the fact that many field studies concentrate on locally derived data only, which may not be generalized. Mathematical modelling of the glacial system has therefore become an important tool in glacial geomorphological studies, especially for the assessment of the possible effects of the many different interacting processes at the ice–bed interface.

It is generally accepted that large Pleistocene ice sheets underwent important subglacial melt due to the insolation effect of geothermal warming and by frictional heating due to ice flow. There are many cited examples where basal meltwater is important in explaining subglacial morphological features with respect to glacier beds of unlithified sediments, such as the deformable bed (Boulton and Jones, 1979), channelized subglacial meltwater transport (Shoemaker, 1986; Alley, 1989), drumlin formation (Boulton, 1987) and

glaciotectonic features (van der Wateren, 1992; Mooers, 1990). Others have pointed out the eroding power of subglacial meltwater and its relation to the formation of tunnel valleys (Nye, 1976; Wright, 1973; Boulton and Hindmarsh, 1987) and, to a certain extent, to the origin of glacial basins (van der Wateren, 1992). We follow the method applied by Shoemaker (1986), Boulton and Jones (1979) and Mooers (1990) for modelling subglacial hydrology, but we put emphasis on longitudinal ice sheet trends.

We consider a two-dimensional longitudinal segment through an ice sheet, parallel to the ice flow direction, consisting of an ice sheet and an aquifer overlying an aquiclude. The initial upper surface of the aquifer is a perfect horizontal plane. Groundwater flow is discharged entirely through this confined aquifer. The hydrological properties of interest are the amount of meltwater and the ability of the aquifer to drain the meltwater by groundwater flow.

Groundwater flow through the aquifer is modelled by Darcy's law:

$$Q = \frac{KD}{\rho_w g}\frac{dP_w}{dx} \tag{1}$$

where Q(m^2s^{-1}) is the amount of water flow through the aquifer, K ($m\,s^{-1}$) is the hydraulic conductivity, D (m) is the thickness of the aquifer, ρ_w ($kg\,m^{-3}$) is the density of water, g ($m\,s^{-2}$) is the acceleration due to gravity, P_w ($kg\,m^{-1}\,s^{-2}$) is the water pressure, and x (m) is the distance measured in the longitudinal direction parallel to the maximum ice sheet topography gradient.

Maximum water pressure in the aquifer can be determined by evaluating the physical properties of the glacier bed due to ice overburden, which is related to the longitudinal topography of the ice sheet, by assuming that the soil of the bed at any place is at the point of failure (Shoemaker, 1986). The effect of soil failure and resulting sediment deformation on water pressures, as outlined by Boulton and Hindmarsh (1987), is neglected in our calculation and only discussed qualitatively, because it would complicate the analysis unnecessarily. The criterion for soil strength is modelled by Coulomb's equation:

$$\tau_b = c + (\rho_i g h - P_w)\tan\phi \tag{2}$$

and the relation to ice topography:

$$\tau_b = \rho_i g h \mathrm{d}h/\mathrm{d}x \tag{3}$$

where τ_b ($kg\,ms^{-1}\,s^{-2}$) is the basal shear stress, c ($kg\,m^{-1}\,s^2$) is soil cohesion (zero for sandy aquifers), h (m) is height, $\rho_i g h$ ($kg\,m^{-1}\,s^2$) is the ice overburden pressure, and $\tan\phi$ is the coefficient of internal friction of the soil. The slope of the ice sheet topography ($\mathrm{d}h/\mathrm{d}x$) is calculated by using a simple empirical equation (Paterson, 1982):

$$h = A\sqrt{x} \tag{4}$$

where A ($m^{1/2}$) is a coefficient, normally having a value between 2·5 and 5 (Paterson, 1982). Topographic profiles of Pleistocene ice sheets are likely to have had low values for A, as can be concluded from the effects of glacier sliding on deformable beds (Boulton and Jones, 1987).

Maximum water pressure is calculated when $\mathrm{d}h/\mathrm{d}x$ in Equation 3 is replaced by the derivative of Equation 4 and combined with Equations 2 and 3. For sandy aquifers ($c = 0\,kg\,m^{-1}\,s^{-2}$) this results in:

$$P_{w(max)} = \rho_i g h - [\{\rho_i g h(A/2\sqrt{x})\}\tan\phi] \tag{5}$$

Darcy's law can be used to calculate the water pressure necessary to allow a given amount of meltwater to be transported entirely through the aquifer. The water pressure needed for complete groundwater drainage is referred to as $P_{w(gw)}$. Assuming that the rate of meltwater production (m) equals the rate of aquifer outflow at the ice sheet margin, $P_{w(gw)}$ is calculated as:

$$P_{w(gw)} = 1/2(m\rho_w g/KD)x^2 \tag{6}$$

Typical values for basal melt rate m_b vary from 0·005 to 0·02 m a^{-1} and depend on geothermal heat flux, heat from deformation friction due to ice flow and sliding over the bed (Hooke, 1977). Instability of the glacial system occurs when calculated water pressures ($P_{w(max)}$) are greater than maximum water pressures allowed

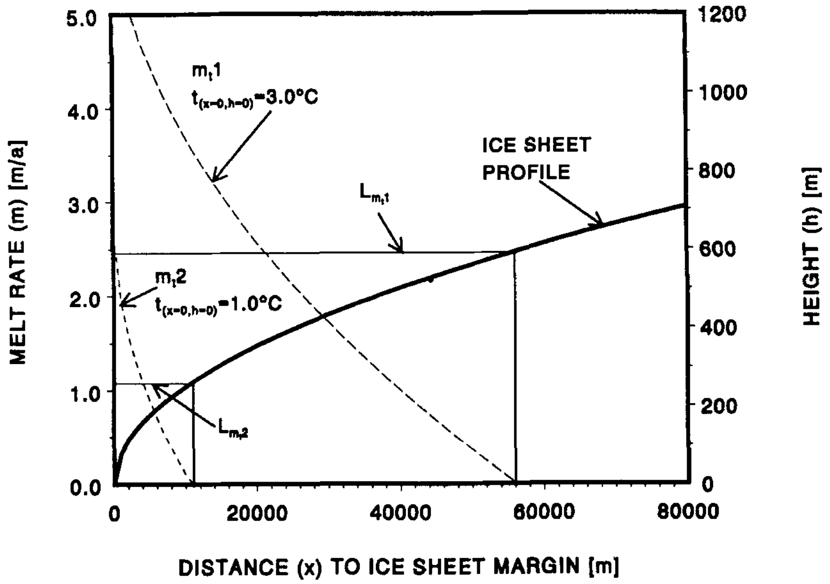


Figure 3. Glaciohydrological longitudinal cross-section parallel to the direction of maximum ice sheet topography gradient, showing different types of aquifer water pressures (in elevation heads H of water equivalents, $H = P/(\rho g)$. The recalculation of pressures to elevation head is done to allow comparison with ice sheet topography. The curve representing elevation head for complete groundwater drainage with constant basal melt rate ($H_{(gw)}, m_b = 0{\cdot}03\,\mathrm{m\,a^{-1}}$) is everywhere below the maximum elevation head $H_{(max)}$. Addition of supraglacial meltwater to the subglacial system leads to an imminent negative effective head, which is $H_{(max)} - H_{(gw)}$

by the soil failure criterion (Equation 2), $P_{w(max)}$ (Figure 3). In reality, this instability will almost never occur, as the subglacial hydrological system will respond by the development of a drainage system at the ice–bed interface, thereby decreasing the difference between water pressures needed for water flow through the aquifer and maximum water pressures in order to prevent soil failure. Boulton and Hindmarsh (1987) considered the existence of subglacial channels due to imminent hydrological instabilities crucial for the development of tunnel valleys. In our analysis, we will support this concept but give an alternative interpretation of the influence of controlling factors.

Modelling the flow of meltwater beneath large Pleistocene ice sheets, as described, is only valid when there is no restriction on groundwater leaving the subglacial aquifer at the ice sheet margin, as it would be as a result of the presence of permafrost. As suggested by van der Wateren (1985), high fluvial activity in front of the ice sheet margin (Gibbard, 1988) probably prevented the development of a frozen toe at the margins of the ice sheet, and is used as an argument in our model to assume free groundwater flow.

Values for K and D in our model are based on characteristics of the sandy pre-Elsterian deposits which reach thicknesses of up to 300 m (Bosch, 1991). Although evidence on deep groundwater movement during glaciation is currently not available, subglacial groundwater modelling experiments by Boulton *et al.* (1993) do indicate the importance of high aquifer thicknesses in The Netherlands. We will therefore study groundwater flow beneath ice sheets with high values for D from Equation 1.

Application of high values for both K and D for modelling groundwater flow causes no unstable groundwater drainage with basal melt rates of $0{\cdot}03\,\mathrm{m\,a^{-1}}$ (Figure 3). The development of a subglacial drainage system in areas of thick unconsolidated sediments with medium hydraulic conductivities, such as expected during the Elsterian glaciation in northwestern Europe, can therefore only be explained by assuming additional meltwater sources.

Supraglacial meltwater

Under climatological conditions of subpolar and temperate ice sheets, supraglacial meltwater becomes important and during summers will generally exceed the amounts produced by subglacial melt. Values of several metres per year are common where ice sheets are temperate. The relation between air temperature

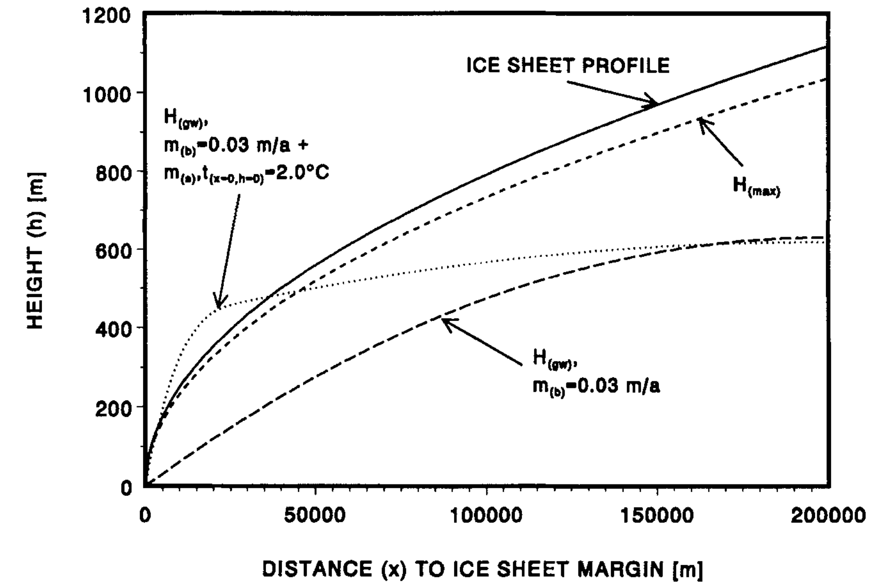


Figure 4. The effect of mean average temperature at the ice sheet margin $t_{(x=0, h=0)}$ on available meltwater at the glacier bed

and melt rate is not straightforward and can only be calculated for our goals by rigorous simplification of glacier climatology. An empirical equation is derived from a set of equations describing the heat balance of ice sheet surfaces (Paterson, 1982, p. 299–312; Kuhn, 1987; Röthlisberger and Lang, 1987), where from each equation meteorological constants have been combined and values for variables have been set to a reasonable average. Supraglacial melt rate m_s is here given by:

$$m_s = rt + e \tag{7}$$

where t is air temperature (°C) above the ice sheet surface, r is a coefficient and e is a constant. Values for r and e depend on the radiation balance (net radiation, sensible heat, latent heat), albedo, wind speed and roughness at the ice surface, atmospheric water vapour pressure, and energy from precipitation. A complete outline of glacier meteorology is not within the scope of our analysis and does not affect the main principles of glacier hydrology. Maximum supraglacial meltwater production is expected at the ice sheet margin and decreases with respect to x, hyperbolic down to zero at the distance (x) where supraglacial melt (m_s) is zero (Figure 4).

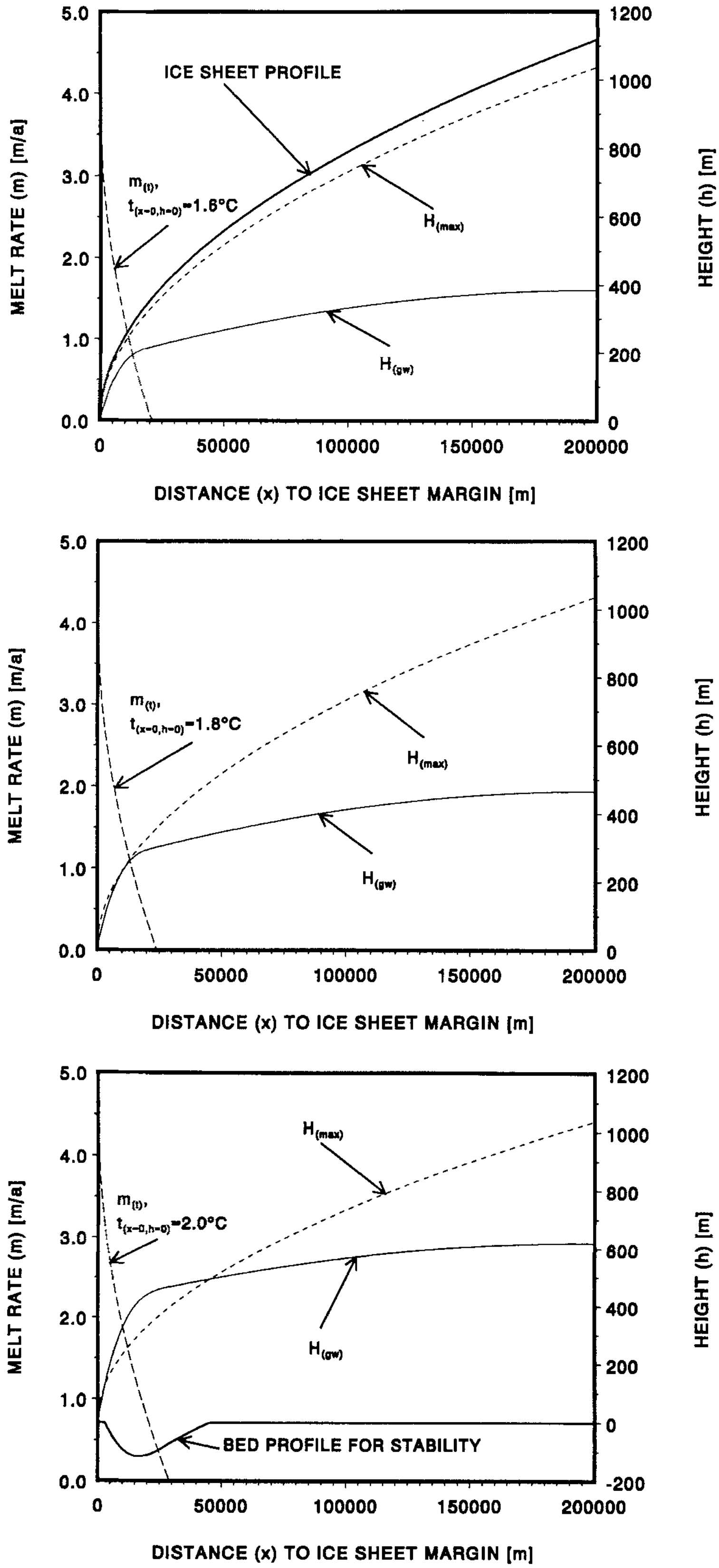
For the longitudinal profile it is necessary to calculate supraglacial melt rates at any point at the glacier surface. The linear relationship between air temperature and height is given by the wet adiabatic temperature gradient of 6°C per 1000 m (0·006°C m^{-1}). If the temperature t at the ice sheet margin ($x = 0, h = 0$) is known, we can calculate the air temperature above the ice sheet surface as a function of h:

$$t = t_{(x=0,h=0)} - 0{\cdot}006\,h \tag{8}$$

Combining Equation 8 with Equation 7 enables the calculation of the longitudinal zone where supraglacial meltwater is produced. Because of the parabolic shape of the ice sheet profile, the effect of changes in coefficient A of Equation 4 and the value for $t_{(x=0,h=0)}$ will be exponential on the supraglacial meltwater-producing zone L_m(m), which is calculated by combining Equations 4, 7 and 8:

$$L_m = \left[\frac{(-e/r) - t_{(x=0,h=0)}}{-0{\cdot}006A}\right]^2 \tag{9}$$

Meltwater produced at the ice sheet surface must reach the glacier bed in order to become important for subglacial hydrology. In the analysis of Shreve (1972) it is demonstrated that ice sheets or glaciers, the main part of which is at a temperature at or slightly above the melting point, are characterized by a


ICE SHEET PROFILE
$m_{(t)}$, $t_{(x=0,h=0)}$=1.6°C
$H_{(max)}$
$H_{(gw)}$
MELT RATE (m) [m/a]
HEIGHT (h) [m]
DISTANCE (x) TO ICE SHEET MARGIN [m]
$m_{(t)}$, $t_{(x=0,h=0)}$=1.8°C
$H_{(max)}$
$H_{(gw)}$
MELT RATE (m) [m/a]
HEIGHT (h) [m]
DISTANCE (x) TO ICE SHEET MARGIN [m]
$H_{(max)}$
$m_{(t)}$, $t_{(x=0,h=0)}$=2.0°C
$H_{(gw)}$
BED PROFILE FOR STABILITY
MELT RATE (m) [m/a]
HEIGHT (h) [m]
DISTANCE (x) TO ICE SHEET MARGIN [m]

relatively open structure for the transport of supraglacially produced meltwater. Surface meltwater reaches the bed through moulins and crevasses at places where ice flow properties allow this. We will assume that supraglacial meltwater is transported directly to the glacier bed at the distance from the ice sheet margin where it is produced. Total melt rate (m_t) becomes:

$$m_t = m_s + m_b \tag{10}$$

and is used for the calculation of $P_{w(gw)}$ in Equation 6 by replacing m with m_t.

Stability of the subglacial drainage system

When high values for extra supraglacial melt rates are considered, even thick, permeable aquifers beneath ice sheets will rapidly reach their maximum transport capacity, causing imminent unstable sliding conditions and eventually the development of subglacial channels. The origin and stability of the subglacial drainage system have been studied extensively. The case of a glacier resting on a hardrock bed was modelled by Röthlisberger (1972) and Nye (1973). Shoemaker (1986), Boulton and Hindmarsh (1987) and Alley (1989) extended this theory to subglacial hydrology beneath ice sheets lying on a bed consisting of unconsolidated sediments.

Since unlithified sediments are more easily eroded than hardrock, it seems very likely that subglacial channels beneath ice sheets resting on beds of unlithfied sediments are of the N-channel type (Nye, 1973). However, the survival chance of a subglacial drainage system developed in unlithified sediments is limited (Alley, 1989). Temporal and spatial variations in meltwater supply, in particular, cause the drainage system to be unstable on longer time scales.

The main problems in applying existing theories on subglacial meltwater drainage through subglacial channels include: (1) the high rate of change of meltwater supply to the glacier bed in the longitudinal direction; (2) the high longitudinal hydraulic gradients which cause decoupling of the aquifer–channel water flow and cause predominantly longitudinal groundwater flow; (3) channel instability caused by sediment creep into the channels by deformation of the glacier bed and blocking of the channel by relatively excessive sediment supply caused by either low channel flow velocity during winters or by high erodibility of the glacier bed upstream of the subglacial channels.

DISCUSSION

Taking into account all possible parameters that could affect the stability of subglacial hydrology, an alternative model for glacial erosion of large ice sheets overlying thick beds of unconsolidated sandy sediments of medium permeability is derived. A numerical example of a hypothetical deglaciation scenario is applied on the model equations, where it is assumed that this deglaciation is achieved by an increase in mean air temperature. Of interest are three steady-state glaciohydrological conditions during deglaciation as a function of increasing temperature (Figure 5, Table I).

When production of supraglacial meltwater increases and becomes important, as expected during deglaciation, water pressures beneath the ice sheet will reach values close to the overburden pressure (Figure 5a). This will first happen at some distance from the ice sheet margin (Figure 5b) and subglacial channels are formed. Sediment deformation is likely to occur simultaneously with subglacial channel formation, but will be of limited importance because of the low cohesion of the sandy sediments. Flow of water within the channels incised into the unconsolidated sediments of the glacier bed will lead to erosion of the channel

Figure 5. Three different glaciohydrological conditions that might have occurred during deglaciation of a Scandinavian ice sheet in northwestern Europe as a function of mean air temperature at the ice sheet margin. Table I gives numerical values used for model parameters based on reasonable averages from the literature. (a) Elevation head for complete groundwater drainage $H_{(gw)}$ with air temperature of 1·6°C at the ice sheet margin. Meltwater discharge is entirely by groundwater flow. (b) Elevation head for complete groundwater drainage $H_{(gw)}$ with air temperature of 1·8°C at the ice sheet margin. The curve $H_{(gw)}$ touches the curve $H_{(max)}$, which means that maximum groundwater flow is established. (c) Elevation head for complete groundwater drainage $H_{(gw)}$ with air temperature of 2·0°C at the ice sheet margin. Subglacial channels are formed and erode the glacier bed to the stable longitudinal profile, as given by the solid line

Table 1. Numerical values for model equation parameters as applied to the graphs in Figures 5a–c and 6

Variable	unit	Equation
$K = 1 \times 10^{-4}$	$\mathrm{m\,a^{-1}}$	1, 6
$D = 300$	m	1, 6
$\rho_w = 1000$	$\mathrm{kg\,m^{-3}}$	1, 6
$\rho_i = 927$	$\mathrm{kg\,m^{-3}}$	2, 3, 5
$g = 9{\cdot}81$	$\mathrm{m\,s^{-2}}$	1, 2, 3, 5, 6
$A = 2{\cdot}5$	$\mathrm{m^{1/2}}$	4, 5, 9
$m_b = 0{\cdot}01$	$\mathrm{m\,a^{-1}}$	10
$r = 1{\cdot}8$		7, 9
$e = 10$		7, 9
$\phi = 35°$		2, 5

beds. This removal of considerable amounts of sediment by erosion within the subglacial channels undeniably results in a lowering of the glacier bed.

During winters, when the supply of supraglacial meltwater is negligible, ice flow or sediment creep will almost certainly fill in the subglacial channels. At the start of the melt season, meltwater flows towards the subglacial channels and causes water pressures to build up. The channel slopes, in particular, are susceptible to glacial kinematic erosion and will cause widening of the cross profile of the subglacial channel under zero or slightly negative effective pressures during the summer (Boulton, 1974, 1979; Boulton *et al.*, 1974). The lowering of the subglacial channel beds by fluvial erosion and widening by glacial kinematic erosion will eventually create an elongated subglacial depression.

Assuming that the longitudinal ice sheet profile retains its form during erosion of subglacial channels, lowering of the glacier bed leads to a local increase in overburden pressure (Figure 5c). The effect of an increase in overburden pressure is the additional strength of the upper layers of the glacier bed, which permits $P_{w(max)}$ and thus the transport capacity of the aquifer to be locally increased. The increase of $P_{w(max)}$ is only expected where $P_{w(gw)} > P_{w(max)}$, as calculated without taking into account channelized drainage. The increase of $P_{w(max)}$, as a result of the eroding bed, continues to both sides from the initial point of glacier bed erosion, until the equilibrium $P_{w(gw)} = P_{w(max)}$ is reached at every point along the longitudinal profile, where $P_{w(gw)} > P_{w(max)}$. The final result will be that the subglacial drainage system is no longer needed and that meltwater drainage is entirely through the aquifer (Figure 5c). This situation can be regarded as a condition of stable groundwater flow.

Stable groundwater flow is established when imminent unstable pressure differences caused by increasing supraglacial meltwater supply are compensated by sufficiently high erosion rates of subglacial channels. Calculations of sediment transport capacity of subglacial meltwater channels, made by de Groot *et al.* (1993), show that erosion rates are probably not the limiting factor for reaching any subglacial depression depth during deglaciation. We therefore assume that an increase in supraglacial meltwater supply, causing imminent unstable glaciohydrological conditions, is instantaneously compensated by a correction of the glacier bed topography due to erosion, establishing stable groundwater flow.

With this approach it is not necessary to actually calculate erosion rates, because once a condition of stable groundwater flow is reached, no further erosion can take place, and erosion can only be initiated again by an increase in temperature of the ice sheet margin (Figure 6). The existence of subglacial channels and thus erosion is only expected during the transition from one stable glaciohydrological condition to the next. It should be stressed that application of the concept of stable groundwater flow conditions only enables the definition of maximum erosion depth and does not refer to erosion rates. Mean air temperatures are calculated to an accuracy of one decimal place, in order to show the high sensitivity of the glaciohydrological system to changes in mean air temperature and,

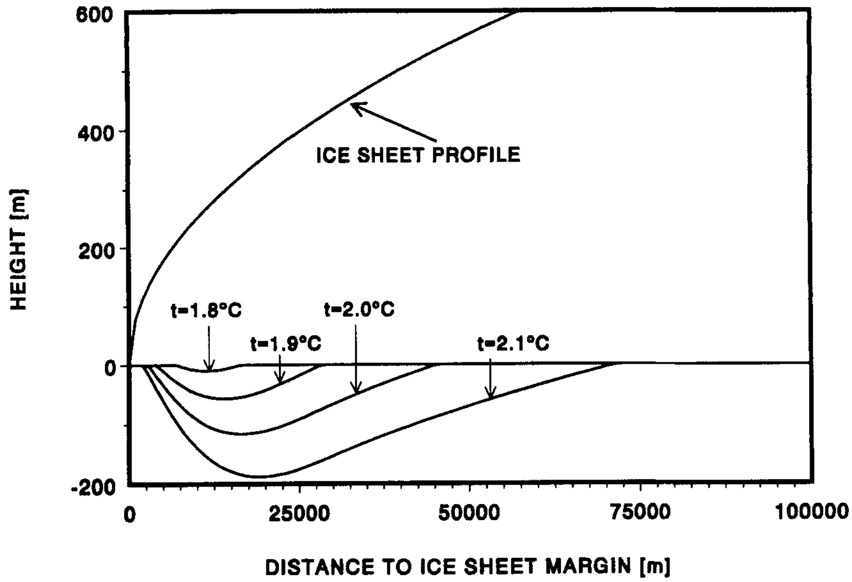


Figure 6. Maximum longitudinal tunnel valley profiles for different air temperatures at the ice sheet margin

because of the lack of data on temperatures during deglaciation of Quaternary ice sheets, they are only indicative.

The formation of tunnel valleys and glacial basins

Because of the relationship between climate and erosion depth in our model (Figure 6), it is proposed that differences in glacial geomorphology of a particular glaciation can be related to differences in climatic conditions during glaciation. This proposition applies only to the main characteristics of glacial morphology. Local phenomena such as drumlins, eskers and small and isolated tunnel valleys in unconsolidated sediments are probably related to special local conditions. We will discuss the consequences of our theory by making a glaciomorphological comparison between the tunnel valleys of the Elsterian and the glacial basin–push moraine of the Saalian as they are explained as a function of specific climatic conditions during deglaciation.

Tunnel valley development during the Elsterian can be interpreted as the result of relatively rapid and continuing increase in ice sheet margin temperatures, accompanying increasing meltwater supplies. High values for meltwater production during deglaciation, which are needed to maintain a progressive retreat, prevent the tunnel valleys from reaching stable groundwater flow conditions. The fact that Elsterian tunnel valleys are rarely found to be disturbed by large glaciotectonic features may support the assumption of a progressively retreating ice sheet margin. This could also explain the relatively close spacing and deep, narrow appearance of most Elsterian tunnel valleys. Phases of rapid retreat alternated with periods of slower retreat. If these changes in retreat velocity are assumed to be related to changing mean summer temperatures, meltwater production during a slow retreat phase was at a significantly lower level compared with the rapid retreat phases. Since meltwater production is directly related to potential erosion depth in our model, this mechanism could account for the many thresholds found at the bases of tunnel valleys. The valley thresholds mark the positions of historic stable ice sheet margins where subglacial erosion rates were low. A locally more constant retreat of the ice sheet margin may have caused longer tunnel valleys with no thresholds.

The landscape in front of the retreating ice sheet margin may have been characterized by the formation of proglacial lakes, rapidly filling with a continuing supply of meltwater. High water levels in these lakes may have led to the uplift of the ice to form a floating and calving ice sheet margin. Sediment infill of these lakes

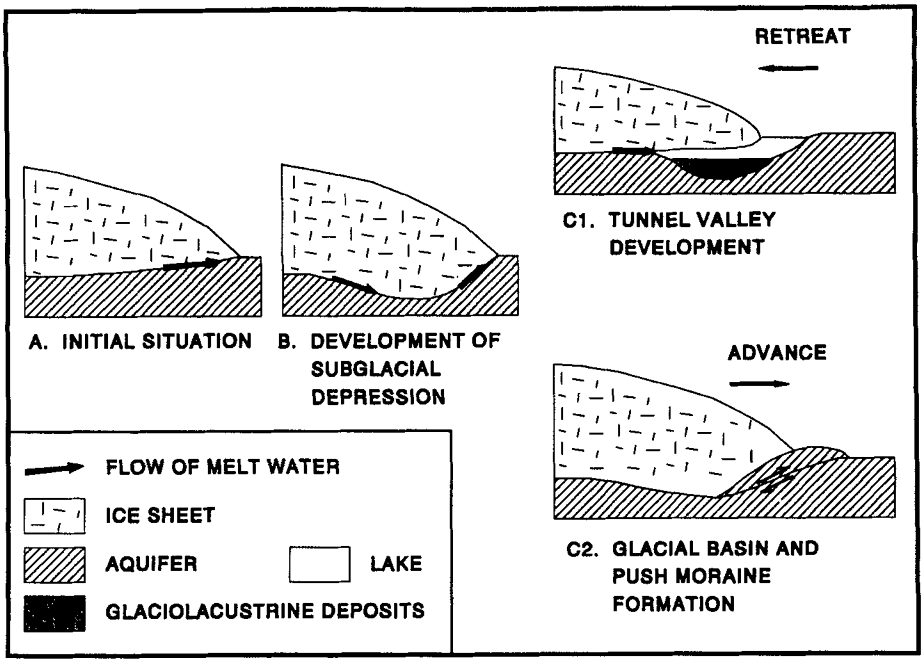


Figure 7. Schematic representation of the possible genetic analogy between glacial basin–push moraine development and the origin of tunnel valleys

consists of the erosion products of active tunnel valleys such as coarse outwash fan deposits and clay-rich lake sediments from the melting floating ice sheet (Figure 7, C1).

Saalian push moraine formation, on the other hand, may be regarded as the effect of a readvancing ice sheet in shallow subglacial depressions. During deglaciation of the Saalian ice sheet, temperature rise may not have been as rapid as during the Elsterian. Depression spacing was relatively large, due to low supraglacial meltwater supply. When mean annual temperatures dropped, the ice sheet readvanced onto the depression slopes to form push moraines (Figure 7, C2), thereby increasing both maximum basin depth and width. Meltwater erosion in subglacial channels at the base of the glacial basins probably continued during push moraine development, which may account for the occurrence of the typical depression geometry resembling those to tunnel valleys at the proximal side of some glacial basins.

CONCLUSIONS

Current models describing the drainage of meltwater beneath a large ice sheet overlying unconsolidated sediments do not predict glacial erosion in thick beds of relatively high permeability. Only by incorporating the effects of supraglacial meltwater production which occurred during deglaciation, is a model based on generally accepted methods for modelling subglacial hydrology of large ice sheets able to explain the occurrence of large-scale subglacial erosive depressions, and even the formation of deep Elsterian tunnel valleys, in northwestern Europe. The theoretical model is based on the assumption that erosion in subglacial channels can be explained as a temporary response to overcome potentially unstable glaciohydrological conditions during periods of excessive meltwater production. Theoretically derived longitudinal profiles of deep glacial depressions resemble tunnel valley geometry as described in the literature. Very deep subglacial depressions can only be formed in easily erodible unconsolidated sediments, and their formation could be well controlled by climatic conditions during deglaciation. The model presented suggests that differences between large-scale glacial morphological landscapes of different glaciations in areas with predominantly unconsolidated sediments can be regarded as the product of a specific deglaciation scenario.

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